RAINING A MAGMA OCEAN: THERMODYNAMICS OF ROCKY PLANETS AFTER GIANT IMPACTS.

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Introduction. Giant impacts partially vaporize rocky planets. The outer layers of the post-impact body are sufficiently hot that volatiles and silicates are dissolved into a single fluid [1-3]. As the body cools, the silicate liquid and volatile gases separate into distinct phases. The initial partitioning of volatiles between the newly-formed magma ocean and atmosphere is determined by the evolution of the pressure-temperature profile through the body coupled to the internal dynamics.

Predictions of the volatile content of a magma ocean are often based on solubility-controlled equilibrium at the magma ocean-atmosphere interface. Because the magma ocean is turbulently convecting, most studies have assumed that magma that was in contact with the atmosphere would efficiently mix with the rest of the magma ocean [e.g., 4]. The volatile budget of the magma ocean just prior to freezing is then based on the solubilities of different species at the pressure-temperature conditions near the liquidus.

However, mantle isotopic variations are evidence against perfect mixing in the last magma ocean and over Earth's history [e.g., 5, 6]. The assumption of efficient mixing is violated when the molten planet is rotating. The interior flow structure in the rotating fluid is likely organized into Taylor columns rather than large convective cells [Fig. 1 & Fig. 17 in 7]. Mixing is inhibited in the radial direction from the center of a vortex [e.g., 8], and mixing is complex along the vertical axis of a column and within the tangent cylinder to the core [7, 9]. We find that the Rossby number, the ratio of inertial to Coriolis forces, is small in magma oceans, typically 0.01 to 1. Consquently, angular momentum is an important factor in the dynamics of Earth's last magma ocean for all proposed models of the Moon-forming giant impact.

In this work, we investigate the thermodynamic path a planet follows while cooling from a vaporized body to a magma ocean near the liquidus. We discuss the implications for the volatile content of the magma ocean. The changes in energy and timescale for cooling are discussed in [10].

Post-impact rocky bodies are supercritical silicate fluids. The post-impact body (a planet or synestia [1]) is strongly thermally stratified. The upper silicate layer has much higher entropy compared to the lower silicate layer. Figure 1 presents the mean specific entropy of the upper 25 wt% of the silicate layer for the post-impact bodies with masses between 0.9 and 1.1 M_{Earth} produced by giant impacts in the database from [1]. The silicate was modeled with the MANEOS forsterite equation of state [11-13]. The majority of giant impacts generate post-impact bodies with outer layers that are mostly vapor with specific entropies greater than the critical point value. Because vapor has a much lower density than liquid silicate, the body is much greater in size compared to the cooled planet [e.g., Fig. 10 in 1]. In most cases, the high-entropy outer layer transitions smoothly from vapor to supercritical fluid with increasing pressure. Thus, the post-impact rocky body has a phase structure that is similar to the gas giant planets and does not have a liquid surface.

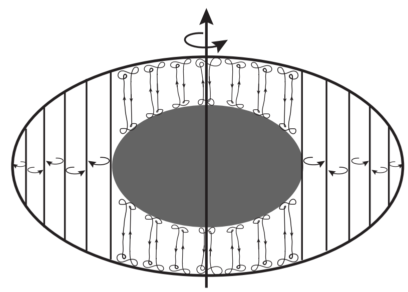

Fig. 1. Schematic cross section showing Taylor columns in a rapidly rotating magma ocean around a metal core [e.g., 9]. Angular momentum conservation inhibits flow perpendicular to the rotation axis.

Cooling the supercritical silicate fluid. In all potential Moon-forming giant impacts, the high-entropy layer is sufficiently hot that volatiles are initially dissolved into the silicate vapor and supercritical fluid of the post-impact body [1-3]. The thermal state of the circumplanetary disk, or disk-like region in a synestia, is more variable. However, the mass of the disk is less than the total mass of high-entropy silicate, and because the disk has a large radiating surface area, it cools more quickly than the planet. The planet cools little in the time it takes for the disk to fully condense. Volatiles accreted by the planet from the disk would be dissolved into the supercritical fluid and vapor.

Earth's volatile budget (<1wt%) is a much smaller mass fraction than the high-entropy silicate layer [2]. For simplicity, we assume that the volatiles were homogeneously dissolved into the high-entropy layer because concentration variations would have been reduced by mixing during the impact. Radiation from the photosphere cools the gas and generates silicate droplets that fall into hotter regions and vaporize. Vigorous convection of the fluid driven by radiative cooling of the high-entropy layer combined with silicate rain cools the body and maintains a quasi-isentropic thermal profile (at least parallel to the spin axis) that transitions to a saturated silicate vapor profile [2].

Initially, the pressure-temperature profile of the high-entropy silicate intersects the vapor side of the liquid-vapor boundary (e.g., the 7 kJ/K/kg isentrope in Fig. 2 intersects at about 200 bar) and then follows the saturated vapor curve to lower pressures. As the body radiatively cools, the mean entropy of the upper layer decreases and the intersection point with the vapor curve increases in pressure (e.g., about 1 kbar at 6 kJ/K/kg in Fig. 2) until it reaches the critical point (25.5 kbar for this forsterite equation of state).

When the upper layer cools just below the critical point entropy, the magma ocean liquid surface forms. At this point, silicate vapor comprises a few wt% of the planet, and most of the volatiles may remain dissolved into the higher pressure silicate fluid. The solubilities of volatiles near the critical point of bulk silicate Earth (BSE) are not known. We are calculating the solubilities of volatiles in BSE composition using first principles molecular dynamics simulations [3]. In general, water and hydrogen solubility increases in silicate liquids with pressure [3, 14, 15]. Therefore, the initial magma ocean may form with a relatively high volatile content.

Magma ocean surface pressure decreases with time. With continued silicate rain, the atmospheric mass decreases, the atmospheric composition becomes enriched in volatiles, and the atmospheric pressure over the magma ocean decreases. When the atmosphere becomes dominated by volatile species, the magma ocean thermal profile cools away from the silicate vapor curve. The atmospheric pressure is lower for high angular momentum bodies compared to non-rotating bodies because of Coriolis forces. At this point, the surface pressure may be substantially lower than usual estimate of 100s bar, and the silicate isentrope will be near the liquidus of the lower mantle, which will begin freezing soon after the surface temperature cools below the silicate vapor curve. Because rotation inhibits complete mixing, the deep magma ocean volatile content may not be reset by pressure and temperature changes at the surface.

Implications for volatile evolution. Initially, the high-entropy layers of the magma ocean contains volatiles that were dissolved during the giant impact. The magma ocean surface forms at the critical point, where water and hydrogen solubilities will be large. As the magma ocean cools, the solubilities of volatiles will change, but the rotating system restricts mixing processes. Thus, the evolution of volatiles in the magma ocean may not be primarily controlled by solubility at the cooling magma ocean-atmosphere interface. Instead, partitioning of volatiles into the magma ocean may be determined by the initial dissolution during the giant impact and partial re-equilibration at the magma ocean-atmosphere interface that is limited by the dynamics of the rotating silicate fluid.

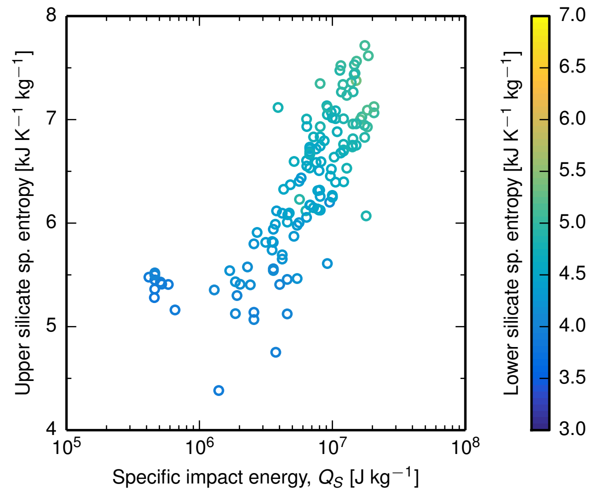

Fig. 2. Giant impacts vaporize the outer layers of post-impact bodies. The mean entropies of the outer 25 wt% of silicates are correlated with the specific impact energy and are generally greater than the critical point value (5.4 kJ/K/kg). Data from [1].

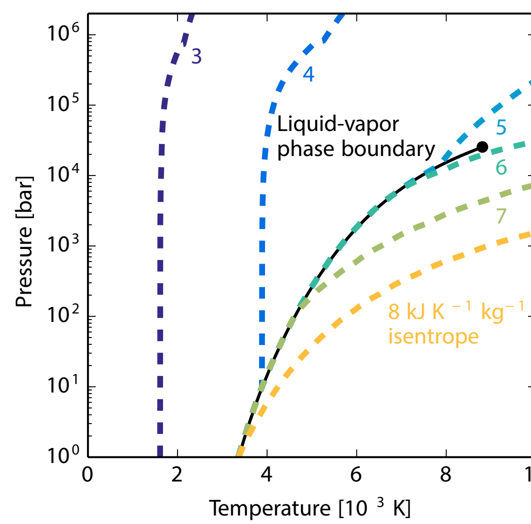

Fig. 3. Pressure-temperature profiles for quasi-isentropic outer layers of post-impact planets using the forsterite MANEOS equation of state.

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